From slow to fast faulting – recent challenges in earthquake fault mechanics

S. Nielsen[1]

Manuscript accepted for publication in PHILOSOPHICAL TRANSACTION A.
Final publication 28 September 2017, available at http://dx.doi.org/10.1098/rsta.2016.0016

[1]Durham University, Earth Sciences Department, Durham DH1 5ED, UK
stefan.nielsen@durham.ac.uk

Abstract

Faults –thin zones of highly localised shear deformation in the Earth– accomodate strain on a momentous range of dimensions (millimetre to hundreds of kilometres for major plate boundaries) and of time intervals (from fractions of seconds during earthquake slip, to years of slow, aseismic slip and millions of years of intermittent activity). Traditionally, brittle faults have been distinguished from shear zones which deform by crystal plasticity (e.g. mylonites). However such brittle/plastic distinction becomes blurred when considering (1) deep earthquakes that happen under conditions of P, T where minerals are clearly in the plastic deformation regime (a clue for seismologists over several decades); (2) the extreme dynamic stress drop occurring during seismic slip acceleration on faults, requiring efficient weakening mechanisms. High strain rates ($> 10^4$ s^{-1}) are accomodated within paper-thin layers (principal slip zone or PSZ), where co-seismic frictional heating triggers non-brittle weakening mechanisms. In addition, (3) pervasive off-fault damage is observed, introducing energy sinks which are not accounted for by traditional frictional models. These challenge our traditional understanding of friction (R & S laws), anelastic deformation (creep and flow of crystalline materials) and the scientific consensus on fault operation.

Faults are active at very different depths (0-30 km for the Earth crust, but up to 700 km for observed intermediate and deep earthquakes). Faults live under conditions of ambient pressure and temperature that cover a very wide span –indicatively, 1-300 MPa and 20-300^{o}C in the Earth crust alone; > 10 GPa and $> 600^o$ for deep earthquakes.

Their range of deformation rates and styles is also very wide. A single fault may undergo slip at both extremely fast (m/s, seismic rates) and extremely slow (cm/y, inter-seismic creep, up to 50 cm/y for some slow slip events).

Different slip rates usually map onto different portions of the fault (locked vs. unlocked fault patches, deep vs. shallow fault areas). Such variety in fault behaviour is attributed to changes in friction, originating in variations of structural or compositional fault properties, temperature, normal stress and presence of fluid pressure.

This volume is mostly focussed on findings about the co-seismic fault deformation –the fast aspect of fault sliding– which is evolving into an extremely active and productive research area. In the following sections I offer an opinion on some recent and less recent findings which are challenging our working hypothesis on earthquake slip. These findings broadly fall in three categories which are reflected in the contributions to this special volume.

1 A fault paradigm

For perhaps the last five decades, the working hypothesis in classic fault research has broadly adopted the following assumptions:

(a) Fault deformation and earthquakes are principally associated to slip (displacement discontinuity) across a surface of negligible thickness (rather often simplified to a planar surface). Deformation accomodated within a larger volume around the fault, other than the purely elastic strain, is generally neglected when considering fault kinematics, dynamics and dissipative terms in the earthquake energy budget. Therefore, it is widely believed that faults are controlled mainly by friction on a surface (or across a gouge layer of millimetric or sub-millimetric thickness).

(b) Velocity-strengthening friction induces stable, slow sliding while velocity-weakening friction is responsible for potentially unstable, fast seismic slip. Such behaviour is often modelled using Rate-and-State friction laws, which are documented by traditional laboratory tests, where slip velocity steps are imposed using high stiffness machines. Although these allow to explore only slow slip rates ($<$ mm/s), it was believed quite intuitively that velocity strengthening observed at such rates would preclude acceleration to seismic rates ($>$ 1 m/s) by preventing the onset of unstable behaviour. Therefore, form the phenomenological point of view, until recently a velocity strengthening material was associated to a stable sliding fault portion and vice-versa, where no seismic slip was to be expected under any circumstance.

(c) From the micro-physical mechanism point of view, brittle regime is characteristic of seismic slip, while plastic or viscous deformation (crystal flow, diffusion creep, viscous shear of melt) mostly occurs in slow deformation processes either diffuse or localised.

A number of observations have come to challenge the above working paradigm.

2 Challenging observations

2.1 Dissipation: is it only friction?

Fault zones comprise one or several localised shear bands which do accommodate most of the slip [1], which have been named Such Principal Slip Zones (PSZ) or principal slip surfaces. The PSZ is generally thin ($50\ \mu\text{m} - 1$ cm), representing high to extreme shear localisation in fault strands which show evidence of having hosted seismic activity at shallow (≈ 1 km) [2–4], intermediate ($\approx 4-10$ km) [5–9] or great depth (upper mantle) [10–12]. Depending on fault-rock composition, the PSZ is variably constituted of ultracataclasis [2, 13, 14], recrystallised nano- to micro-grains [8, 15, 16], or quenched melt [3, 4, 17, 18, and references therein]. In more mature faults the PSZ is surrounded by a fault core (typically a few decimeters) of cataclastic material. (Note that often times, especially for minor fault strands, the PSZ and the fault core are assimilated). A wider damage zone (typically a few meters, on relatively mature faults) with pervasive diffuse damage in the form of microcracks [19] secondary fault veins [20] or pulverised rock [21, 22] is observed surrounding the fault core. Distributed damage has been attributed to the stress concentration around the rupture tip [23], to Mach fronts during supershear ruptures [24], or to stress fluctuations associated to fault roughness at the large or small scale [see 25, and references therein].

Friction laws derived from models or laboratory experiments consider only slip on a fault surface (or within a thin PSZ). Therefore many earthquake models based on fault friction alone implicitly neglect off-fault dissipation. Cowie and Scholtz [26] observed from field data that the size of the breakdown zone scales with the length of the fault, therefore that energy loss per unit fault area should also scale with fault length. Additional laboratory experiments [23, 27] and field studies on natural faults [20, 28–31] also indicated that the width of damage zone increases with fracture length. As pointed out by Nielsen [32, 33] this offers an interpretation for the apparent discrepancy between fracture energy in large earthquakes (estimated from seismology), and fracture energy resulting from frictional weakening under seismic slip conditions (measured in laboratory experiments). Both fracture energies are compatible from small to moderate slip amounts ($\Delta u \leq 0.3$ m), but appear to diverge for large slip and large earthquake magnitudes, where a larger ratio of off-fault dissipation to frictional work is expected.

Mechanical work is the product of stress and strain, and because most anelastic fault strain is accommodated by slip within the PSZ, off-fault damage in itself does not necessarily indicate a large amount of dissipated energy. On another hand, it can be argued that work involved with off-fault damage can be significant if sliding dynamic friction is low, as I will argue in the short discussion below.

In terms of mechanical work, dissipation per unit fault area due to anelastic strain ν_{ij} can be written as:

$$W = \int_H dz \int_{\gamma_{ij}} \tau_{ij}(\nu_{ij})\ d\nu_{ij} \tag{1}$$

(assuming implicit summation on repeated indexes) where H is the thickness of fault zone and z the fault-normal direction. Defining a representative average for σ_{ij} throughout the deformation episode with γ_{ij} as the final strain, we may write:

$$\sigma_{ij} \equiv \frac{1}{\gamma_{ij}} \int\limits_{\gamma_{ij}} \tau_{ij}(\nu_{ij})\ d\nu_{ij} \tag{2}$$

the above reduces to:

$$W = \int\limits_{H} \sigma_{ij}\gamma_{ij}\ dz \tag{3}$$

Now splitting in the contribution of the PSZ and the rest of the fault zone we may write:

$$W = \int\limits_{h} \gamma_{ij}\sigma_{ij}\ dz + \int\limits_{H'} \gamma'_{ij}\sigma'_{ij}\ dz \tag{4}$$

where h is the thickness of the PSZ and $H' = H - h$ the remaining fault zone, and the prime indicates values outside the PSZ. Taking average values of strain and stress within h and H' yields:

$$W = h\ \gamma_{ij}\sigma_{ij} + H'\ \gamma'_{ij}\sigma'_{ij} \tag{5}$$

Strain in the PSZ is dominated by the fault-parallel shear $\frac{\partial u_x}{\partial z}$ (i.e. fault slip), where z is fault normal direction and x the direction of slip; therefore we can write the finite shear strain strain as:

$$\begin{aligned} \gamma_{zx} = \gamma_{xz} = \frac{1}{2}\left(-\frac{\partial u_x}{\partial z}\cancel{\frac{\partial u_x}{\partial x}} - \cancel{\frac{\partial u_z}{\partial z}}\cancel{\frac{\partial u_z}{\partial x}} + \cancel{\frac{\partial u_z}{\partial x}} + \frac{\partial u_x}{\partial z} \right) \\ \approx \frac{1}{2}\frac{\partial u_x}{\partial z} \\ \approx \frac{1}{2}\frac{\Delta u}{h} \end{aligned} \tag{6}$$

where Δu is fault slip, and additional strain terms in the PSZ ($\gamma_{xx}, \gamma_{yz}, \ldots$) are negligible.

Outside the PSZ, for an indicative bulk anelastic deformation $\Delta u'_i$ (note that here deformation is intended as a displacement difference, a opposed to dimensionless strain), over a characteristic dimension H' we may write (neglecting quadratic terms):

$$\gamma'_{ij} = \frac{1}{2}\left(\frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i}\right) \approx \frac{1}{2}\left(\frac{\Delta u'_i}{H'} + \frac{\Delta u'_j}{H'}\right). \tag{7}$$

Then implicitly summing all possible i, j in $\gamma'_{ij}\sigma'_{ij}$ for the off-fault mechanical work, we may write using (5-6):

$$W' = \frac{1}{2}\left(\Delta u'_i + \Delta u'_j\right)\sigma'_{ij} \tag{8}$$

hence summing all terms and using stress symmetry we can re-write the anelastic strain dissipation as:

$$\begin{aligned} W &= \frac{1}{2}\Delta u\ \sigma_{xz} + \frac{1}{2}\Delta u\ \sigma_{zx} + \frac{1}{2}\left(\Delta u'_i + \Delta u'_j\right)\sigma'_{ij} \\ &= \Delta u\ \sigma_{xz} + \frac{1}{2}\left(\Delta u'_i + \Delta u'_j\right)\sigma'_{ij}. \end{aligned} \tag{9}$$

In the particular case that the fault-parallel component is dominant in the off-fault deformation $\Delta u'$, this may be approximated by:

$$W \approx \Delta u\ \sigma_{xz} + \Delta u'\sigma'_{xz} \tag{10}$$

and reverting to the integral form (substituting σ_{ij} and Δu according to their values in 2 and 6) we obtain:

$$W = \int\limits_{u} \tau_{ij}(u)\ du + \int\limits_{u'} \tau'_{ij}(u')\ du' \tag{11}$$

Strikingly, this simplified formulation of (10) and (11) shows that mechanical work, and its ratio inside vs. outside the PSZ, does not depend on h, H but solely on deformations $\Delta u, \Delta u'$ and stress on- and off-fault. This aspect may facilitate field measurements with the aim to estimate dominant off-fault deformation, by simply summing slip on individual fault strands. The reason for not equating σ'_{xz} and σ_{xz} above is that both deformations are not necessary simultaneous therefore may take place under different stress level, as further discussed below. This expression will hold whether $\Delta u'$ results from diffuse deformation or from the sum of localised slip on a number of fractures; continuity of traction allows to use significant average stress throughout H. I note, for consistency, that expression (11) for the case of fault-parallel anelastic shear alone is compatible with the more rigourously derived result of the classic J-integral applied to shear faulting [34], but (11) allows to decouple on- and off-fault contributions, whether they are localised or diffuse.

Because all terms of stress are bounded by the material strength, and that $\Delta u \gg \Delta u'$, we may expect $W \approx \Delta u\, \sigma_{xz}$ and a negligible contribution to mechanical work from off-fault deformation $\Delta u'$. However, most of the slip Δu takes place when the fault friction is extremely low owing to dynamic weakening. (During seismic slip $\sigma_{xz} = \mu\, \sigma_n$ i.e., frictional shear stress under normal stress $\sigma_n = \sigma_{zz}$; experiments report friction coefficient as low as $\mu = 0.05$ under co-seismic slip conditions). On the other hand, stress during the rupture initiation is high and, during that initial phase in the rupture process, fault slip is still small ($\Delta u \approx 0$).

Therefore it is conceivable that under mostly low sliding friction, the off-fault term W' is not negligible in W. The contribution to dissipation due to deformation in the larger volume around the fault will be significant due the initial phase of rupture where the proximity of the rupture tip induces modest deformation $\Delta u'$ but large stress concentrations, resulting in significant product of both, while Δu is still small or comparable to $\Delta u'$. During later phases of rupture the term $\Delta u\, \sigma_{xz}$ may increase only relatively because σ_{xy} drops to extremely low frictional values. Note that W is not the equivalent fracture energy G, which is obtained by simply subtracting the relaxed, sliding shear stress value τ_r from the stress in the W expression:

$$\begin{aligned} G = \int_u (\tau_{xz}(u) - \tau_r)\ du + \int_{u'} (\tau'_{xz}(u') - \tau_r)\ du' \\ = W - \tau_r(\Delta u + \Delta u') \end{aligned} \tag{12}$$

Therefore, using only slip on the PSZ and frictional evolution predicted for a single fault strand (left hand integral in equation 12) would result in an underestimate of the fracture energy.

As a consequence of self-similarity in fracture mechanics, fault slip Δu and off-fault deformation $\Delta u'$ increase proportionally to rupture length. Slip weakening means that friction becomes less significant with on-going slip and rupture growth, while the relative importance of dissipation from off-fault damage continues to increase. Therefore larger earthquakes should be dominated by off-fault dissipation, while small earthquakes are still dominated by friction.

The importance of off-fault damage in the energy budget, and its expected increase with earthquake magnitude (or rupture length) given the above arguments, has also been investigated using numerical simulations which allow for ductile deformation off-fault when material strength is exceeded. To my knowledge the first numerical study on this aspect was contributed by Andrews [35], where a Coulomb yield criterion was applied to the maximum resolved shear stress in any orientation within the volume surrounding the fault surface. At each numerical time step, any stress in excess of the Coulomb limit was compensated by allowing an equivalent amount of anelastic shear strain, and the corresponding anelastic work was computed. With the set of parameters used in Andrew's study, off-fault dissipation initially exceeded on-fault dissipation when rupture length surpassed about 300 m, thereafter increasing linearly as a function of rupture length. More sophisticated models were subsequently proposed [36–42] confirming the findings of [35]. Other studies instead include explicit fault splays branching from the main fault surface [43, 44] as a form of broadening the damage zone.

In addition to mechanical work dissipated by anelastic deformation, energy is also spent in the creation of new surface during rupture process. The amount of surface on the main fault only is insignificant, but does not remain such when adding a multitude of sub-faults and microcracks in the vicinity of the fault. A significant amount of new surface can be obtained only by generating a very dense network of micro-cracks, in particular, if comminution and thin pulverisation involves a significant volume of the fault zone with creation of sub-millimeter fragments (it is readily shown that the amount of surface in a given volume of fragmented material increases as the inverse radius of the fragments). An extreme form of such pulverisation is observed on a number of seismic fault outcrops [21, 24, 45]. No significant deformation is observed within those pulverised rocks in the vicinity of the fault core (the minerals and structures are still recognisable) but the cohesion of the rock has vanished due to the small scale fragmentation with very high microcrack density. One interpretation is that pulverisation develops under extremely high strain rates which are induced nearby the fault in the vicinity of the propagating the rupture tip; in particular, in the case of super-shear rupture propagation episodes, strain-rate is intensified within the Mach-cone sheet of the shear wave radiated from the rupture tip region [24].

Pulversiation under high strain rates has been reproduced in the laboratory using Hopkinson impact bars to create short pulses of intense stress on rock samples [24, 46–48]. In a recent example of such experimental work, Barber and Griffith [49, this volume] argue that the surface energy represents a substantial proportion of the total mechanical energy under extreme loading conditions, possibly an energy sink comparable to the amount of frictional work on a seismic fault.

Off-fault damage (due to anelastic strain, pulverisation or both) provides a mechanism for stopping large earthquakes. As strain energy release rate scales with rupture length, a larger and larger fracture energy (or equivalent energy sink) is required to limit rupture propagation velocity and eventually to arrest the earthquake rupture. Because frictional dissipation remains bounded (to avoid the paradox of negative friction [33]), and because friction all but vanishes in the dynamic sliding steady-state [50], larger earthquakes would never stop unless off-fault dissipation is significant. In this situation earthquake faults –which are essentially shear cracks– become similar to mode I cracks (opening or tension cracks) where friction is absent and fracture energy is controlled by ductile deformation in a finite volume around the crack tip.

The above observations challenge the assumption (a): that the energy sinks in the earthquake rupture process are dominated by friction.

2.2 Fault slip: stable, unstable, or both?

Faults accommodate slip in a variety of fashions: by relatively constant slow, stable sliding; during episodic slow slip events which sometimes generate tremor; by seismic slip at high slip velocity (m/s) and fast rupture propagation (km/s) accompanied by radiation of elastic waves. Such spectrum of behaviours can be viewed as going from the most stable to the most unstable condition.

Potential to generate instability and eventual seismic rupture depends on the two criteria that:

(1) the steady-state frictional stress τ_{ss} on the fault decreases with slip U and slip velocity V (velocity weakening behaviour):

$$\frac{\partial \tau_{ss}}{\partial U} < 0 \quad \text{and} \quad \frac{\partial \tau_{ss}}{\partial V} < 0 \tag{13}$$

(2) the frictional relaxation is faster than the stress release due to slip. In the latter criterion, fault stiffness and a critical frictional stiffness are compared and the condition for instability may be written [51] as the inequality:

$$C\frac{\mu'}{L} < -\frac{V}{D_c}\,\frac{\partial \tau_{ss}}{\partial V} + f(V, \dots) \tag{14}$$

where the left hand side represents fault stiffness (the ratio of shear modulus μ' to the fault dimension L, times a dimensionless geometrical factor C). The right hand side represents the velocity dependence of friction, with additional terms $f(.)$ involving inertia and the state of evolution of the fault. Because fault stiffness is positive, criterion (14) implicitly relies on (13), therefore (13) is necessary but not sufficient condition for instability.

Traditionally the frictional behaviour of rocks has been determined through experiments conducted at slow sliding velocities (microns to millimeters per second) –these velocities are in excess of the tectonic loading rates ($\simeq$ cm/y), but still orders of magnitude lower than during co-seismic slip (m/s). Based on these experimental results, a mathematical form of friction known as Rate-and-State (R & S) law [52], with one or more evolving state variables, can be defined. With a single state variable, such friction is governed by two dimensionless parameters a and b (representing direct effect upon a velocity step and the subsequent evolution effect, respectively) and an evolution distance Dc.

The difference $(b - a)$ represents the dependence of friction on logarithm of velocity, therefore determining the velocity dependence in (13). Then under R & S friction and $a - b < 0$ it can be shown (assuming negligible inertia and evolution term such that $f(V,\dots)=0$), that that the critical frictional stiffness in the right-hand side of (14) becomes:

$$k_c = -\frac{V}{D_c}\frac{\partial \tau_{ss}}{\partial V} = \frac{\sigma_n(b-a)}{D_c} \tag{15}$$

where σ_n is normal stress. The condition (14) with (15) is very widely used as criterion to predict stability of fault materials. There are complications in this stability criterion when rupture is allowed to propagate ($L \neq$ const). In such case it can be shown numerically that both $(b-a)$ and a/b will control the onset of instability [53].

In the limit case where $C\frac{\mu'}{L} \approx \frac{\delta\tau}{D_c}$, (or $a \approx b$ under R & S law), it is believed that small fluctuations in properties combined with episodic slow slip can generate emergent, small-scale instability resulting in tremor-like, prolonged behaviour almost below the limit of instrumental detection, and in the frequency range between 1 and 10 Hz. This may happen in the transition region between the deep, stable and the shallower, unstable portion of major plate boundary faults [54].

The general expectation has been that fault patches which show steady slip behaviour (as observed from geodetic measurements) were constituted by material with velocity-strengthening properties which would remain such throughout the seismic cycle. Given the range of velocities (and duration) where R & S friction was characterised and the emphasis on $(b-a)$ as a control parameter, slip stability criteria have often been applied to natural faults by extrapolating laboratory data of several orders of magnitude toward the lower (inter-seismic) and the higher (seismic) regimes. The measured value of $(b-a)$ in many materials constituents of fault rocks is very small, and the velocity dependence in R & S laws is logarithmic; therefore the predicted weakening upon extrapolation from laboratory to seismic velocities is modest (usually a few percent).

However, extreme dynamic weakening of friction is now being very widely observed under fast (seismic) slip rates. Such knowledge has been established only recently, because the technical implementation of high-velocity, seismic like conditions in laboratory experiments has been achieved no earlier than the '80 with the pioneering work of Shimamoto and co-workers [55] and became widespread in the last decade [see 8, 33, 56–72, among others]. It is notable that such high-velocity weakening has been documented even on rocks which show rate-strengthening in traditional, slow frictional tests. These comprise clay-like material which is expected in the accretionary prism traversed by the shallowest part (repeatedly interpreted as stable sliding) of seismogenic oceanic thrust faults [73].

An empirical formulation that combines the traditional R & S results with an enhanced dynamic weakening as the inverse of velocity was been proposed by Zheng and Rice [74] who used it in numerical modelling of seismic slip pulses. The experimental foundation for such an empirical formulation was subsequently demonstrated by Spagnuolo [72] based on high velocity, rotary shear experiments on slica-built and carbonate-built cohesive rocks. Such formulation can be synthetised as follows:

$$\mu_{ss} = \frac{\mu_o + (a-b)\log(V/V_o)}{1+(V/V_c)^p} \tag{16}$$

where μ_{ss} is the steady-state friction coefficient. The top part of the fraction represents the usual R& S formulation of friction, where μ_o and V_o reference values for friction and sliding velocity, respectively. The denominator, on the other hand, is introduced to account for substantial weakening to kick-in when sliding velocity V is close to, or larger, than a characteristic velocity V_c. The experimental data on both high and low velocity friction were best fit by using $0.08 < V_c < 0.13$ m/s, and an exponent p indicatively in the range $0.4-1$ (the higher values were obtained on carbonatic rocks), in combination with usual a, b parameters for R & S friction.

To illustrate the effect of the velocity dependence in (16) on fault stability, we can compute the corresponding critical frictional stiffness (again assuming that we are close to the steady-state and that inertia is negligible). Taking the indicative value $p=1$ in (16) we obtain:

$$k'_c = \frac{\sigma_n}{D_c}\left(\frac{b-a}{\frac{V}{V_c}+1} + \frac{\mu_o + (a-b)\log(\frac{V}{V_o})}{\left(\frac{V}{V_c}+\frac{V_c}{V}+2\right)}\right) \tag{17}$$

where we retrieve k_c as of (15) in the limit $V \ll V_c$, but here the critical stiffness varies and peaks substantially in the vicinity of $V = V_c$, as argued in [72]. This indicates that fault instability is enhanced if the experimentally observed velocity weakening is allowed to kick-in.

It is striking that in (16) even if $(a - b)$ is positive, velocity weakening can be achieved at sub-seismic slip rates, and k'_c will also become positive and peak close to $V = V_c$. Therefore a slight acceleration may allow the fault to become unstable.

However, a slight acceleration on a creeping fault section means that the fault instability somehow has already started –a chicken and egg situation of sorts– therefore the extended criterion (17) cannot be applied to the slow nucleation phase, but would assume some independent triggering process. An example of such triggering was illustrated in a numerical model [75] where a velocity-weakening, unstable fault patch generated the initial instability. Seismic slip then propagated into an otherwise stable creeping patch, making it dynamically weak (here the assumed mechanism for weakening was thermal pressurization of the fluids during fast slip on the creeping patch).

To corroborate these laboratory experimental and these numerical results, co-seismic slip has been observed within the shallow portion of thrust faults, which are clay-rich regions of typical velocity-strengthening behaviour. The most striking example is the large co-seismic slip observed during the Tohoku, 2009 earthquake in the up-dip part of the trench. Models based on a more traditional set of assumptions had failed to forecast coseismic slip in such portion of the fault [76] and had to be subsequently revisited [77]. In this case the interpretation is that the shallow portion of the fault gradually accelerated under the impulse of rupture propagating from larger depths of the fault, until it too reached a critical weakening velocity. Additionally, during the so-called tsunami earthquakes, ruptures appears to initiate within the shallow portion thrust faults, an even more paradoxical situation in terms of the expectation of stable-sliding in shallow trench faults.

In response to the recent Tohoku, 2011 earthquake, Noda [78, this volume] propose new ways to incorporate the enhanced dynamic weakening in a numerical model of thrust fault behaviour. Contrary to previous attempts [74], Noda chooses to model the enhanced weakening by incorporating a quadratic law for log velocity dependence in friction. They fit the empirical law using data from laboratory friction experiments, which were performed on samples collected from the Japan trench during the J-FAST drilling project following the earthquake. According to their model of earthquake cycle on a vertical section of the fault, essentially two types of earthquakes can occur: large catastrophic events which break through the topmost fault section to reach the ocean bottom, or intermediate events which are confined at depth to the blueshist region.

Finally, recent laboratory experiments provided evidence that velocity-hardening friction does not necessarily preclude the spontaneous nucleation ot seismic-like, stick-slip instability if a sufficient level of small-scale inhomogeneity is introduced on the fault [79]. The experiments were performed on cohesive, pre-cut samples of Westerly granite, where simulated faults had been prepared by grinding the surface to achieve variable levels of roughness, and submitted to variable levels of confinement (from 30 to 200 MPa). The experimental set-up (direct shear of cohesive, pre-cut rocks) shares similarities with earlier experiments [80, 81], but with a much higher confinement stress and smaller scale, as in more recent realisations [82, 83]. In [79], velocity-stepping was imposed during the experiments, allowing to measure the dimensionless parameters a, b and the weakening distance D_c characterising R & S friction. Under a sub-set of conditions, a clearly positive value of $a-b$ was found, which indicates rate-hardening and is classically believed to generate only stable slip. Even within this very subset indicating $a - b > 0$, stick-slip and seismic instability episodes were triggered inducing partial melt of the slip surface. Using scaling arguments and the self-affine roughness of natural faults, the authors proposed that the same mechanism may trigger stick-slip instability on natural earthquake faults. The interpretation of this unstable behaviour, is that when the inhomogeneity the fault is enhanced (in this case by roughness), weak patches are formed due to normal stress fluctuations which act as stress concentrators and initiators of instability.

As reported in the experimental study proposed by Rutter and Hackston [84, this volume], a stable-sliding fault can also become seismic under the effect of a fluid injection. Here, too, we note that the effect of inhomogeneity is crucial, as rupture is triggered when the rate of injection is high, preventing pressure diffusion to a larger zone and creating a localised region of low effective normal stress. Fluid pressure can locally increase due to natural causes –the failure of a nearby seal of underground volatiles [85]– or to anthropogenic activity [86] such as injections to enhance hydrocarbure or hydrothermal productivity.

These observations challenge the stability concept (b): that fault portions showing stable creep –or within material that shows stable R & S behaviour under slow slip conditions– cannot undergo seismic slip or the nucleation of stick-slip instability.

2.3 Seismic shear deformation: brittle, plastic, viscous... all of the above?

Deformation associated to crustal faults is typically described as a combination of dry frictional sliding on a surface (or within gouge in a PSZ of minimal thickenss), surrounded by a fault zone of intense fracturing and cataclasis; all such processes typically belong to the brittle regime. However, growing evidence of crystal plastic deformation and melting has been reported both on natural and experimentally simulated seismic faults. I discuss below the implications of these, together with other deformation and weakening mechanisms essentially triggered by an abrupt co-seismic temperature rise.

A classic problem arising when considering earthquake slip at velocities in excess of 1 m/s, occurring under the typical shear stress levels expected from dry friction (see former paragraph, R & S laws), is the intense heating and temperature rise which should occur on the fault. It is well known [17, 87] that melting temperatures of the rock should be reached even for moderate size earthquakes by a simple back-of-envelope calculation for a thickness h of the PSZ and thermal diffusion of a heat rate τV:

$$\begin{aligned} \Delta T &= \frac{1}{\rho c\sqrt{\kappa\pi}}\mu\sigma_n V\sqrt{t} \qquad \text{for } h \ll 2\sqrt{\kappa t} \\ \Delta T &= \frac{1}{\rho c h}\mu\sigma_n V t \qquad \text{for } h \gg 2\sqrt{\kappa t}. \end{aligned} \tag{18}$$

(with typical values for crystalline rock of $\kappa = 1.6\ 10^{-6}$ m^2/s, $\rho = 3500$ kg/m^3, $c = 1000$ J kg^{-1}K^{-1}). Assuming a static friction angle of $\theta = 30^o$, and Andersonian stress state, the value of normal stress would be 3/4 the lithostatic load for a fault near static failure (Sibson, 1974). Further assuming hydrostatic pore pressure at a depth of 10 km, an indicative value of effective normal stress would result in $\sigma_n \approx 180$ MPa. Under the modest weakening expected for R & S we would still have $\mu \approx 0.56$ and for an earthquake with $V = 1$ m/s, $t = 1$ s we obtain $\Delta T \approx 12800^o$ in the first case of (18) and $\Delta T \approx 2280^o$ assuming adiabatic shear heating within a PSZ of $h = 10^{-2}$ m in the second. In both cases some severe alteration of the fault conditions is expected in the early phases of seismic slip, with phase transitions (melting, decomposition, amorphisation, dehydration, decarbonation in the case of carbonatic rocks such as limestone or dolostone), supercritical fluid pressurization, and the thermal triggering of efficient shear strain mechanisms which therefore would mitigate further rise in temperature (as indeed no significant heat flow anomaly is detected near active faults [88, and references therein]. These arguments have fostered an extremely active research field regarding the thermally triggered weakening mechanisms likely to take place on earthquake faults.

Two of these mechanisms are re-visited by Rice [89, this volume]: flash weakening, a process which is first encountered in early metallurgic research [90] and which can be mathematically formalised and extended to slip on fault rocks [91]; and thermal pressurization of either native fluids trapped in the fault zone or volatiles resulting from co-seismic decomposition reactions. Both models have been extensively used in earthquake rupture modelling [75, 92, and references therein]. While flash-weakening is widely observed under fast slip conditions, direct experimental evidence of thermal fluid pressurization is scarce. The first unequivocal evidence of thermal pressurization has been documented in two experimental studies by by Violay et al. [70, 93], but they show that pressurization should start after several meters of slip only. Alternative weakening mechanisms are more efficient in the initial phases of slip which kick-in at very earlier slip stages buffering the background temperature rise and initially preventing an efficient pressurisation. For the types of simulated faults used in [70, 93] and the extrapolation they offer to natural faults, it can be argued that pressurization is more likely a mechanism arising during large earthquakes, providing further weakening to a fault which has already achieved lubrification.

Frictional melting, on the other hand, appears as a rather intuitive consequence of co-seismic heating and represents an attractive model for dynamic fault lubrication. Its fossil product, a crypto-crystalline to vitreous solidified melt (pseudotachylyte or PST) is indeed observed on some crystalline fault sections which have been exhumed from intermediate crustal depths [10, 17, 94] or upper mantle shear zones [12] with frequent lateral injection veins.

It is straightforward to obtain melting of crystalline fault rocks during experiments conducted at co-seismic conditions of normal stress and slip velocity [58, 68]. The frictional behaviour observed shows indeed pronounced lubrication under high ($\approx$ 1m/s) slip velocity, where a continuous layer of superheated melt is created that supports viscous shear. The melt layer remains generally thin

($\approx 30\ \mu$m $- 1$ mm) as excess melt is extruded to lateral veins on natural faults, or from the edges of the sample of experimental faults. Theoretical arguments indicate a modest increase of steady-state sliding shear stress τ_{ss} with normal stress σ_n, such that $\sigma \propto \sigma_n^{1/4}$ [63], a trend which is confirmed by experimental investigations in High Velocity Rotary Shear experiments [68, 95]. On natural fossil faults the average co-seismic sliding shear stress can be estimated based on the volume of melt produced [17, 58, 65]; this confirms extremely low (< 0.1) equivalent friction coefficients during dynamic sliding, compatibly with the experimental results.

However, rocks undergoing fast slip show an initial weakening phase that takes place much earlier than background melting temperature is reached, and even before profuse melting is formed. This indicates that the temperature rise and the weakening occur much faster at isolated contact asperities on the sliding surface, due to their large localised stress concentrations –a behaviour clearly indicative of the aforementioned flash weakening.

While profuse melting will start to form only after finite slip (centimeters at typical mid-crustal conditions), unambiguous traces of melting are observed even for minimal slip amounts (fractions of millimetres) when conducting microstructural analysis of crystalline rocks [96, 97] and even under wet conditions [70] in post-experimental samples that slipped under seismic or micro-seismic conditions.

It is also argued that melting is not very generally observed on seismic fault outcrops. There are several possible explanations for the causes of such paucity. First, the melting process may be a rarity which occurs only under a specific set of conditions (immature faults and dry crystalline rocks in continental crust environment). However, it can be rebutted that many of the accessible outcrops expose only the part of the fault which has been seismically active at shallow depth, where normal stress and frictional heating are not sufficient to produce melting, but which are not representative in the budget of mechanical energy release contributing to the rupture. Fossil seismic faults which have been exhumed from depth are more rare. PST veins are often extremely thin and difficult to observe and recognise in the field, unless expected and specifically investigated. Also, the amorphous material constituting PST is easily altered, recrystallised (for example transformed into chlorite or epidote), overprinted or destroyed [98]. Strikingly, products of co-seismic melting and pseudotachylite were observed on samples which were drilled from active faults in the months after an earthquake [3, 4], although the depth was relatively modest (a few hundred meters).

Lastly, on carbonate rock composed predominantly of calcite or dolomite (limestones, marbles or dolostones which host many crustal earthquakes in the Mediterranean area, among others) shear strain appears to localise within a thin ($\approx 50 - 100\ \mu$m) PSZ layer with extreme grain reduction and crystal plastic deformation take place, but no melting. Indeed, thermal dissociation is reached in carbonate rocks at quite lower temperatures than melting, as agued in [59, 60, 99]. In the advanced stages of slip, the PSZ exhibits a densely stacked polygonal aggregate of small (few tens to hundreds of nanometres) crystal grains, with structure typical of grain-boundary sliding plasticity, a regime where superplastic behaviour (the capacity to accommodate finite plastic strain under high deformation rate) has been reported for ceramics. Such structures have been observed in experimental simulated faults after sliding at seismic slip velocity [8, 16] but also on samples of natural faults from tectonic areas involving carbonate rocks [7, 16].

Plastic flow laws generally depend on grain-size D and on temperature T; strain-rate $\dot{\gamma}$ and shear stress τ can be equated by :

$$A\ \tau^n = \dot{\gamma}\ e^{\frac{H}{RT}}\ D^b \tag{19}$$

where H is creep activation energy and A a dimensional normalisation factor. In the case of grain boundary sliding the exponents are $n = 1$ and $2 < b < 3$, with slightly increasing values if dislocation creep component is present [8]. Within the validity domain of (19), observing that the value of $b > 1$ with $D \ll 1$ and the exponential decay with temperature, we may expect strain accomodation to be ever more efficient as temperature increases and grain size decreases. However, this type of flow law has seldom been explored at strain rates in excess of 1 s^{-1}, while in the case for the co-seismic slip in carbonate PSZs the inferred values can be of the order of $\dot{\gamma} = 10^4$ s^{-1}. Therefore, while extreme weakening is observed in combination with dynamic recrystallisation within the co-seismic PSZ, deformation takes place in an uncharted, high-end territory of strain rates, most probably involving flow laws and mechanisms previously unreported.

Plastic flow has also been invoked as the mechanism for deep- and intermediate-focus earthquakes, within silicate-build constituents of the subducting slab, on the double Wadati-Benioff seismicity planes. In this case it is proposed that the superplastic behavior is triggered by phase-transformation, for example Olivine-Spinel transition under pressure increase or dehydration embrittlement of serpentine.

A review of experimental evidence of the latter phase transformation in connection with seismic slip is proposed by Green [100, this volume], who also discuss broader implication for earthquakes. As pointed out by Green, nanometric material weakening has been an intensely debated topic, but remains to date poorly understood. Hypothesis on its possible origin have been proposed such as the inverse Hall-Petch effect; whereas grain-size reduction normally hardens the material by inhibiting dislocation creep, it can enhance other weakening processes such as the grain boundary sliding as explored in [8] and briefly discussed in the preceding paragraph. In any case a growing body of evidence supports the idea that plastic- and melt-related flow does take place on some, if not all, earthquake faults.

These observations challenge assumption (c): that seismic slip should belong to purely brittle regime rather than plastic or viscous regimes.

3 Conclusions

What emerges is a considerably more complex picture of fault behaviour and on the controls of stable versus unstable slip, than that offered by the paradigm of section (1). It will be a non trivial endeavour for the Earth science community, or indeed for a multi-disciplinary science community, to explore the variety of micro-physical mechanisms responsible for the enhanced dynamic weakening which is maintained under at high slip velocity, but also to highlight the way that the weakening mechanisms are triggered in the first place. A quick glance on the some of the ongoing research can be gathered in the present volume on *slow to fast faulting*.

Acknowledgements

I acknowledge the support of Royal Society for publishing this volume and for the help and support in the organisation and hosting of the related conference. I am indebted to the many reviewers who helped to improve the manuscripts included in this volume. I acknowledge R.E. Holdsworth for encouraging me with the conference, and T. Mitchell and A. Schubnel for co-chairing it.

References

[1] Sibson RH. 2003 Thickness of the seismic slip zone. *Bulletin of the Seismological Society of America* **93**, 1169–1178. DOI:10.1785/0120020061

[2] Chester FM, Chester JS. 1998 Ultracataclasite structure and friction processes of the punchbowl fault, san andreas system, california. *Tectonophysics* **295**, 199 – 221.

[3] Boullier AM, Ohtani T, Fujimoto K, Ito H, Dubois M. 2001 Fluid inclusions in pseudotachylytes from the nojima fault, japan. *Journal of Geophysical Research: Solid Earth* **106**, 21965–21977. DOI:10.1029/2000JB000043

[4] Boullier AM. 2011 *Geology of the Earthquake Source: A Volume in Honour of Rick Sibson, Vol. 359, 2011,0.*, volume 359, chapter Fault-zone geology: lessons from drilling through the Nojima and Chelungpu faults, pp. 17–37. Geological Society, London, Special Publication.

[5] Di Toro G, Pennacchioni G, Teza G. 2005 Can Pseudotachylyte be used to infer earthquake source parameters? An example of limitations in the study if exhumed faults. *Tectonophysics* **402**, 3–20. DOI:10.1016/j.tecto.2004.10.014

[6] Fondriest M, Smith SA, Candela T, Nielsen SB, Mair K, Di Toro G. 2013 Mirror-like faults and power dissipation during earthquakes. *Geology* **41**, 1175–1178. DOI:10.1130/G34641.1

[7] Siman-Tov S, Aharonov E, Sagy A, Emmanuel S. 2013 Nanograins form carbonate fault mirrors. *Geology* **41**, 703–706. DOI:10.1130/G34087.1

[8] De Paola N, Holdsworth RE, Viti C, Collettini C, Bullock R. 2015 Can grain size sensitive flow lubricate faults during the initial stages of earthquake propagation? *Earth Planet. Sci. Lett.* **431**, 48–58. DOI:10.1016/j.epsl.2015.09.002

[9] Demurtas M, Fondriest M, Balsamo F, Clemenzi L, Storti F, Bistacchi A, Di Toro G. 2016 Structure of a normal seismogenic fault zone in carbonates: The Vado di Corno Fault, Campo Imperatore, Central Apennines (Italy). *Journal of Structural Geology* **90**, 185–206. DOI:10.1016/j.jsg.2016.08.004

[10] Swanson MT. 1988 Pseudotachylyte-bearing strike-slip duplex structures in the fort foster brittle zone, s. maine. *Journal of Structural Geology* **10**, 813 – 828. DOI:10.1016/0191-8141(88)90097-1

[11] Austrheim H, Boundy TM. 1994 Pseudotachylytes generated during seismic faulting and eclogitization of the deep crust. *Science* **265**, 82–83. DOI:10.1126/science.265.5168.82

[12] Ueda T, Obata M, Di Toro G, Kanagawa K, Ozawa K. 2008 Mantle earthquakes frozen in mylonitized ultramafic pseudotachylytes of spinel-lherzolite facies. *Geology* **36(8)**, 607–610. DOI:10.1130/G24739A.1

[13] Faulkner D, Jackson C, Lunn R, Schlische R, Shipton Z, Wibberley C, Withjack M. 2010 A review of recent developments concerning the structure, mechanics and fluid flow properties of fault zones. *Journal of Structural Geology* **32**, 1557 – 1575. DOI:10.1016/j.jsg.2010.06.009. Fault Zones

[14] Smith SAF, Billi A, Toro GD, Spiess R. 2011 Principal slip zones in limestone: Microstructural characterization and implications for the seismic cycle (tre monti fault, central apennines, italy). *Pure and Applied Geophysics* **168**, 2365–2393. DOI:10.1007/s00024-011-0267-5

[15] De Paola N, Collettini C, Faulkner DR, Trippetta F. 2008 Fault zone architecture and deformation processes within evaporitic rocks in the upper crust. *Tectonics* **27**, TC4017. DOI:10.1029/2007tc002230

[16] Smith S, Di Toro G, Kim S, Ree JH, Nielsen S, Billi A, Spiess R. 2013 Coseismic recrystallization during shallow earthquake slip. *Geology* **41**, 63–66. DOI:10.1130/G33588.1

[17] Sibson R. 1975 Generation of pseudotachylyte by ancient seismic faulting. *Geophys. J. R. Astron. Soc.* **43**, 775–794.

[18] Di Toro G, Pennacchoni G, Nielsen S. 2009 Pseudotachylytes and earthquake source mechanics. In *Fault zone properties and earthquake rupture dynamics* (ed. E Fukuyama), volume 94 of *International Geophysics Series*, chapter 5, pp. 87–133. Elsevier.

[19] Mitchell T, Faulkner D. 2009 The nature and origin of off-fault damage surrounding strike-slip fault zones with a wide range of displacements: A field study from the atacama fault system, northern chile. *Journal of Structural Geology* **31**, 802 – 816. DOI:10.1016/j.jsg.2009.05.002

[20] Savage HM, Brodsky EE. 2011 Collateral damage: Evolution with displacement of fracture distribution and secondary fault strands in fault damage zones. *Journal of Geophysical Research: Solid Earth* **116**, n/a–n/a. DOI:10.1029/2010JB007665. B03405

[21] Dor O, Ben-Zion Y, Rockwell TK, Brune J. 2006 Pulverized rocks in the mojave section of the san andreas fault zone. *Earth Planet. Sci. Lett.* **245**, 642–654. DOI:10.1016/j.epsl.2006.03.034

[22] Mitchell T, Ben-Zion Y, Shimamoto T. 2011 Pulverized fault rocks and damage asymmetry along the arima-takatsuki tectonic line, japan. *Earth and Planetary Science Letters* **308**, 284 – 297. DOI:10.1016/j.epsl.2011.04.023

[23] Moore D, Lockner D. 1995 The role of microcracking in shear-fracture propagation in granite. *Journal of Structural Geology* **17**, 95 – 114. DOI:10.1016/0191-8141(94)E0018-T

[24] Doan M, Gary G. 2009 Rock pulverisation at high strain rate near the san andreas fault. *Nature Geoscience* **2**, 709–712.

[25] Griffith WA, Nielsen S, Di Toro G, Smith FAS. 2010 Rough faults, distributed weakening, and off-fault deformation. *J. Geophys. Res.* **115**, B08409. DOI:10.1029/2009JB006925

[26] Cowie PA, Scholz CH. 1992 Growth of faults by accumulation of seismic slip. *J. Geophys. Res.* **97**, 11,085–11,095.

[27] Zang A, Wagner FC, Stanchits S, Janssen C, Dresen G. 2000 Fracture process zone in granite. *Journal of Geophysical Research: Solid Earth* **105**, 23651–23661. DOI:10.1029/2000JB900239

[28] Vermilye JM, Scholz CH. 1998 The process zone: A microstructural view of fault growth. *Journal of Geophysical Research: Solid Earth* **103**, 12223–12237. DOI:10.1029/98JB00957

[29] Shipton Z, Cowie P. 2001 Damage zone and slip-surface evolution over μm to km scales in high-porosity navajo sandstone, utah. *Journal of Structural Geology* **23**, 1825 – 1844. DOI:10.1016/S0191-8141(01)00035-9

[30] Shipton ZK, Evans JP, Abercrombie RE, Brodsky EE. 2006 The missing sinks: Slip localization in faults, damage zones, and the seismic energy budget. In *Earthquakes: Radiated energy and the physics of faulting: American Geophysical Union, Geophysical Monograph Series 170*, pp. 217–222. DOI:10.1029/170GM22

[31] de Joussineau G, Aydin A. 2007 The evolution of the damage zone with fault growth in sandstone and its multiscale characteristics. *Journal of Geophysical Research: Solid Earth* **112**, n/a–n/a. DOI:10.1029/2006JB004711. B12401

[32] Nielsen S, Spagnuolo E, Violay M, Smith S, Di Toro G, Bistacchi A. 2016 G: Fracture energy, friction and dissipation in earthquakes. *Journal of Seismology* pp. 1–19. DOI:10.1007/s10950-016-9560-1

[33] Nielsen S, Spagnuolo E, Smith SAF, Violay M, Di Toro G, Bistacchi A. 2016 Scaling in natural and laboratory earthquakes. *Geophysical Research Letters* **43**, 1504–1510. DOI:10.1002/2015GL067490. 2015GL067490

[34] Palmer A, Rice J. 1973 The growth of slip surfaces in the progressive failure of over-consolidated clay. *Proc. Roy. Soc. Lond.* **332**, 527–548.

[35] Andrews DJ. 2005 Rupture dynamics with energy loss outside the slip zone. *J. Geophys. Res. (Solid Earth)* **110**, JB003191. DOI:10.1029/2004JB003191. B01307

[36] Ben-Zion Y, Shi Z. 2005 Dynamic rupture on a material interface with spontaneous generation of plastic strain in the bulk. *Earth and Planetary Science Letters* **236**, 486 – 496. DOI:http://dx.doi.org/10.1016/j.epsl.2005.03.025

[37] Bhat H, Dmowska R, King G, Klinger Y, Rice JZ. 2007 Off-fault damage patterns due to supershear ruptures with application to the 2001 mw 8.1 kokoxili (kunlun) tibet earthquake. *J. Geophys. Res.* **112**, 0. DOI:10.1029/2006JB004425

[38] Hok S, Campillo M, Cotton F, Favreau P, Ionescu I. 2010 Off-fault plasticity favors the arrest of dynamic ruptures on strength heterogeneity: Two-dimensional cases. *Geophysical Research Letters* **37**, L02306. DOI:10.1029/2009GL041888. L02306

[39] Gabriel AA, Ampuero JP, Dalguer LA, Mai PM. 2013 Source properties of dynamic rupture pulses with off-fault plasticity. *Journal of Geophysical Research: Solid Earth* **118**, 4117–4126. DOI:10.1002/jgrb.50213

[40] Johri M, Dunham EM, Zoback MD, Fang Z. 2014 Predicting fault damage zones by modeling dynamic rupture propagation and comparison with field observations. *Journal of Geophysical Research: Solid Earth* **119**, 1251–1272. DOI:10.1002/2013JB010335

[41] Xu S, Ben-Zion Y, Ampuero JP, Lyakhovsky V. 2015 Dynamic ruptures on a frictional interface with off-fault brittle damage: Feedback mechanisms and effects on slip and near-fault motion. *Pure and Applied Geophysics* **172**, 1243–1267. DOI:10.1007/s00024-014-0923-7

[42] Viesca RC, Templeton EL, Rice JR. 2008 Off-fault plasticity and earthquake rupture dynamics: 2. effects of fluid saturation. *Journal of Geophysical Research: Solid Earth* **113**. DOI:10.1029/2007JB005530. B09307

[43] Dalguer LA, Irikura K, Riera JD. 2003 Simulation of tensile crack generation by three-dimensional dynamic shear rupture propagation during an earthquake. *Journal of Geophysical Research: Solid Earth* **108**, n/a–n/a. DOI:10.1029/2001JB001738. 2144

[44] Rice JR, Sammis CG, Parsons R. 2005 Off-fault secondary failure induced by a dynamic slip pulse. *Bulletin of the Seismological Society of America* **95**, 109–134. DOI:10.1785/0120030166

[45] Mitchell T, Ben-Zion Y, Shimamoto T. 2011 Pulverized fault rocks and damage asymmetry along the arima-takatsuki tectonic line, Japan. *Earth Planet. Sci. Lett.* **308**, 284–297. DOI:10.1016/j.epsl.2011.04.023

[46] Yuan F, Prakash V, Tullis T. 2011 Origin of pulverized rocks during earthquake fault rupture. *Journal of Geophysical Research: Solid Earth* **116**, n/a–n/a. DOI:10.1029/2010JB007721. B06309

[47] Aben FM, Doan ML, Mitchell TM, Toussaint R, Reuschl T, Fondriest M, Gratier JP, Renard F. 2016 Dynamic fracturing by successive coseismic loadings leads to pulverization in active fault zones. *Journal of Geophysical Research: Solid Earth* **121**, 2338–2360. DOI:10.1002/2015JB012542. 2015JB012542

[48] Zwiessler R, Kenkmann T, Poelchau MH, Nau S, Hess S. 2017 On the use of a split hopkinson pressure bar in structural geology: High strain rate deformation of seeberger sandstone and carrara marble under uniaxial compression. *Journal of Structural Geology* **97**, 225–236. DOI:https://doi.org/10.1016/j.jsg.2017.03.007

[49] Barber T, Griffith WA. 2017 Experimental constraints on dynamic fragmentation as a dissipative process during seismic slip. *Phi. Trans. A* .

[50] Di Toro G, Han R, Hirose T, De Paola N, Nielsen S, Mizoguchi K, Ferri F, Cocco M, Shimamoto T. 2011 Fault lubrication during earthquakes. *Nature* **471**, 494–499. DOI:10.1038/nature09838

[51] Rice JR, Ruina AL. 1983 Stability of steady frictional slipping. *J. Appl. Mech.* **50**, 343–349. DOI:10.1115/1.3167042

[52] Marone C. 1998 Laboratory-derived friction laws and their application to seismic faulting. *Annual Review of Earth and Planetary Sciences* **26**, 643–696. DOI:10.1146/annurev.earth.26.1.643

[53] Rubin AM, Ampuero JP. 2005 Earthquake nucleation on (aging) rate and state faults. *J. Geophys. Res.* **110**, B11312. DOI:10.1029/2005JB003686

[54] Beroza G, Ide S. 2011 Slow earthquakes and nonvolcanic tremor. *Annual Review of Earth and Planetary Sciences* pp. 271–296. DOI:10.1146/annurev-earth-040809-152531

[55] Shimamoto T, Tsutsumi A. 1994 A new rotary-shear high-velocity frictional testing machine: Its basic design and scope of research. *J. Tectonic Res. Group of Japan* **39**, 65–78. (in Japanese with English abstract)

[56] Hirose T, Shimamoto T. 2005 Growth of molten zone as a mechanicsm of slip weakening of simulated faults in gabbro during frictional melting. *J. Geophys. Res.* **110**, B05202. DOI:10.1029/2004JB003207

[57] Hirose T, Shimamoto T. 2005 Slip-weakening distance of faults during frictional melting as inferred from experimental and natural pseudotachylytes. *Bull. Seismol. Soc. Am.* **95**, 1666–1673. DOI:10.1785/0120040131

[58] Di Toro G, Hirose T, Nielsen S, Pennacchioni G, Shimamoto T. 2006 Natural and experimental evidence of melt lubrication of faults during earthquakes. *Science* **311**, 647–649. DOI:10.1126/science.1121012

[59] Han R, Shimamoto T, Hirose T, Ree JH, Ando Ji. 2007 Ultralow friction of carbonate faults caused by thermal decomposition. *Science* **316**, 878–881. DOI:10.1126/science.1139763

[60] Han R, Shimamoto T, Ando Ji, Ree JH. 2007 Seismic slip record in carbonate-bearing fault zones: An insight from high-velocity friction experiments on siderite gouge. *Geology* **35**, 1131. DOI:10.1130/G24106A.1

[61] Mizoguchi K, Hirose T, Shimamoto T, Fukuyama E. 2007 Reconstruction of seismic faulting by high-velocity friction experiments: An example of the 1995 kobe earthquake. *Geophys. Res. Lett.* **34**, L01308. DOI:10.1029/2006GL027931

[62] Tanikawa W, Mishima T, Hirono T, Lin W, Shimamoto T, Soh W, Song SR. 2007 High magnetic susceptibility produced in high-velocity frictional tests on core samples from the chelungpu fault in taiwan. *Geophys. Res. Lett.* **34**, L15304. DOI:10.1029/2007GL030783

[63] Nielsen S, Di Toro G, Hirose T, Shimamoto T. 2008 Frictional melt and seismic slip. *J. Geophys. Res.* **113**, B01308. DOI:10.1029/2007JB005122

[64] Mizoguchi K, Hirose T, Shimamoto T, Fukuyama E. 2009 High-velocity frictional behavior and microstructure evolution of fault gouge obtained from nojima fault, southwest japan. *Tectonophysics* **471**, 285–296. DOI:10.1016/j.tecto.2009.02.033

[65] Nielsen S, Mosca P, Giberti G, Di Toro G, Hirose T, Shimamoto T. 2010 On the transient behavior of frictional melt during seismic slip. *J. Geophys. Res.* **115**, B10301. DOI:10.1029/2009JB007020

[66] Nielsen S, Di Toro G, Griffith WA. 2010 Friction and roughness of a melting rock surface. *Geophys. J. Int.* **182**, 299–310. DOI:10.1111/j.1365-246X.2010.04607.x

[67] Violay M, Nielsen S, Cinti D, Spagnuolo E, Di Toro G, , Smith S. 2011 Friction of marble under seismic deformation conditions in the presence of fluid. In *AGU 2011 Fall Meeting*.

[68] Niemeijer A, Di Toro G, Nielsen S, Di Felice F. 2011 Frictional melting of gabbro under extreme experimental conditions of normal stress, acceleration and sliding velocity. *J. Geophys. Res.* **116**, B07404. DOI:10.1029/2010JB008181

[69] Chang JC, Lockner DA, Reches Z. 2012 Rapid acceleration leads to rapid weakening in earthquake-like laboratory experiments. *Science* **338**, 101–105. DOI:10.1126/science.1221195

[70] Violay M, Nielsen S, Gibert B, Spagnuolo E, Cavallo A, Azais P, Vinciguerra S, Di Toro G. 2013 Effect of water on the frictional behavior of cohesive rocks during earthquakes. *Geology* **42**, 27–30. DOI:10.1130/G34916.1

[71] Smith S, Nielsen S, Di Toro G. 2015 Strain localization and the onset of dynamic weakening in calcite fault gouge. *Earth Planet. Sci. Lett.* **413**, 25–36. DOI:10.1016/j.epsl.2014.12.043

[72] Spagnuolo E, Nielsen S, Violay M, Di Toro G. 2016 An empirically based steady state friction law and implications for fault stability. *Geophysical Research Letters* **43**, 3263–3271. DOI:10.1002/2016GL067881. 2016GL067881

[73] Faulkner DR, Mitchell TM, Behnsen J, Hirose T, Shimamoto T. 2011 Stuck in the mud? earthquake nucleation and propagation through accretionary forearcs. *Geophys. Res. Lett.* **38**, L18303. DOI:10.1029/2011GL048552

[74] Zheng G, Rice J. 1998 Conditions under which velocity weakening friction allows a self-healing versus a cracklike mode of rupture. *Bull. Seismol. Soc. Am.* **88**, 1466–1483.

[75] Noda H, Lapusta N. 2013 Stable creeping fault segments can become destructive as a result of dynamic weakening. *Nature* **493**, 518–521. DOI:10.1038/nature11703

[76] Loveless JP, Meade BJ. 2010 Geodetic imaging of plate motions, slip rates, and partitioning of deformation in japan. *J. Geophys. Res.* **115**, B02410. DOI:10.1029/2008JB006248

[77] Loveless JP, Meade BJ. 2011 Spatial correlation of interseismic coupling and coseismic rupture extent of the 2011 m w= 9.0 tohoku-oki earthquake. *Geophys. Res. Lett.* **38**, L17306. DOI:10.1029/2011GL048561

[78] Noda H. 2017 Earthquake sequence simulations with measured properties for jfast core samples. *Phi. Trans. A* .

[79] Harbord C, Nielsen S, Paola ND, Holdsworth R. 2017 Earthquake nucleation on rough faults. *Geology, in press* .

[80] Okubo PG, Dieterich J. 1984 Effects of physical fault properties on frictional instabilities produced on simulated faults. *J. Geophys. Res.* **89**, 5817–5827.

[81] Ohnaka M, Kuwahara Y, Yamamoto K. 1987 Constitutive relations between dynamic physical parameters near a tip of the propagating slip zone during stick-slip shear failure. *Tectonophysics* **144**, 109–125. DOI:10.1016/0040-1951(87)90011-4

[82] Passelègue FX, Schubnel A, Nielsen S, Bhat HS, Madariaga R. 2013 From sub-rayleigh to supershear ruptures during stick-slip experiments on crustal rocks. *Science* **340**, 1208–1211. DOI:10.1126/science.1235637

[83] McLaskey GC, Lockner DA. 2014 Preslip and cascade processes initiating laboratory stick slip. *J. Geophys Res. (Solid Earth)* **119**, 6323-6336. DOI:10.1002/2014jb011220

[84] Rutter E, Hackston A. 2017 On the effective stress law for rock-on-rock frictional sliding, and fault slip triggered by means of fluid injection. *Phi. Trans. A* .

[85] Miller SA, Collettini C, Chiaraluce L, Cocco M, Barchi M, Kaus BJP. 2004 Aftershocks driven by a high-pressure co2 source at depth. *Nature* **427**, 724–727.

[86] Ellsworth WL. 2013 Injection-induced earthquakes. *Science* **341**. DOI:10.1126/science.1225942

[87] Lachenbruch AH, Sass JH. 1980 Heat-flow and energetics of the san-andreas fault zone. *J. Geophys. Res.* **85**, 6185–6222. DOI:10.1029/JB085iB11p06185

[88] Fulton PM, Brodsky EE, Kano Y, Mori J, Chester F, Ishikawa T, Harris RN, Lin W, Eguchi N, Toczko S, from Expeditions 343, 43T and KR13-08 S. 2013 Low coseismic friction on the tohoku-oki fault determined from temperature measurements. *Science* **342**, 1214–1217. DOI:10.1126/science.1243641

[89] Rice J. 2017 Heating, weakening and shear localization in earthquake rupture. *Phi. Trans. A* .

[90] Archard J. 1959 The temperature of rubbing surfaces. *Wear* **2**, 438 – 455. DOI:10.1016/0043-1648(59)90159-0

[91] Rice JR. 2006 Heating and weakening of faults during earthquake slip. *J. Geophys. Res.* **111**, B05311. DOI:10.1029/2005JB004006

[92] Noda H, Dunham EM, Rice JR. 2009 Earthquake ruptures with thermal weakening and the operation of major faults at low overall stress levels. *J. Geophys. Res.-solid Earth* **114**, B07302. DOI:10.1029/2008JB006143

[93] Violay M, Nielsen S, Spagnuolo E, Cinti D, Di Toro G, Stefano GD. 2013 Pore fluid in experimental calcite-bearing faults: Abrupt weakening and geochemical signature of co-seismic processes. *Earth Planet. Sci. Lett.* **361**, 74 – 84. DOI:10.1016/j.epsl.2012.11.021

[94] Di Toro G, Hirose T, Nielsen S, Shimamoto T. 2006 Relating high-velocity rock-friction experiments to coseismic slip in the presence of melts. In *Earthquakes: Radiated Energy and the Physics of Faulting* (ed. R Abercrombie, A McGarr, H Kanamori, GD Toro), volume 170 of *Geophysical Monograph Series*, pp. 121–134. American Geophysical Union, Washington, D.C. DOI:10.1029/170GM13

[95] Violay M, Di Toro G, Gibert B, Nielsen S, Spagnuolo E, Del Gaudio P, Azais P, Scarlato PG. 2014 Effect of glass on the frictional behavior of basalts at seismic slip rates. *Geophysical Research Letters* **41**, 348–355. DOI:10.1002/2013gl058601

[96] Passelègue FX, Schubnel A, Nielsen S, Bhat HS, Deldicque D, Madariaga R. 2016 Dynamic rupture processes inferred from laboratory micro-earthquakes. *J. Geophys. Res.: Solid Earth* **121**. DOI:10.1002/2015jb012694

[97] Lockner DA, Kilgore BD, Beeler NM, Moore DE. 2017 *The Transition From Frictional Sliding to Shear Melting in Laboratory Stick-Slip Experiments*, pp. 103–131. John Wiley & Sons, Inc. DOI:10.1002/9781119156895.ch6

[98] Kirkpatrick JD, Rowe CD. 2013 Disappearing ink: How pseudotachylytes are lost from the rock record. *Journal of Structural Geology* **52**, 183 – 198. DOI:http://dx.doi.org/10.1016/j.jsg.2013.03.003

[99] Rowe CD, ke Fagereng, Miller JA, Mapani B. 2012 Signature of coseismic decarbonation in dolomitic fault rocks of the naukluft thrust, namibia. *Earth and Planetary Science Letters* **333**, 200 – 210. DOI:http://dx.doi.org/10.1016/j.epsl.2012.04.030

[100] Green H. 2017 Phase-transformation-induced nanometric lubrication of earthquake sliding. *Phi. Trans. A* .